12 THE FUTURE: DRYLANDS

Introduction

Even though arid lands cover one-third of Earth's land surface, remarkably little thought has been given to what might happen to drylands as a result of any potential warming associated with the enhanced greenhouse effect. Indeed, those sections of the reports of the Intergovernmental Panel on Climate Change (IPCC) that deal with deserts are notable for being narrow and for neglecting almost all the most important issues (Bullock and Le Houérou, 1996; Noble and Gitay, 1996). Some of the regional studies are no more satisfactory, and the one on Africa contains one paragraph only on deserts and concludes with the following debatable remark (Zinyowera et al., 1998: 43):

> Extreme desert systems already experience wide fluctuations in rainfall and are adapted to coping with sequences of extreme conditions. Initial changes associated with climate change are less likely to create conditions significantly outside present ranges of tolerance; desert biota show very specialized adaptations to aridity and heat, such as obtaining their moisture from fog or dew.

The reality is that arid environments often appear to have been prone to rapid geomorphologic and hydrologic changes in response to apparently modest climatic stimuli, switching speedily from one state to another when a particular threshold is reached (Goudie, 1994).

1 Desert valley bottoms appear to have been subject to dramatic alternations of cut-and-fill during the course of the Holocene and a large literature has developed on the arroyos (see Chapter 6) of the American southwest (see, e.g., Cooke and Reeves, 1976; Balling and Wells, 1990). Schumm has proposed that valley bottom trenching may occur when a critical valley slope gradient is crossed for a drainage area of a particular areal extent (Schumm, 1977) but other examples result from changes in rainfall intensity and grazing pressure.
2 The heads of alluvial fans often display fan head trenching, suggesting recent channel instability. There has been considerable debate in the literature about the trigger for such trenching, be it tectonic, extreme flood events, climatic change, or an inherent consequence of fluctuating sediment–discharge

relationships during the course of a depositional cycle (e.g., Harvey, 1989).

3 Fluvial systems, as, for example, in the drier lands around the Mediterranean Basin, display a suite of terraces which demonstrate a complex record of cut-and-fill during the late Holocene (Van Andel et al., 1990). There has been prolonged debate as to whether the driving force has been climatic change (Vita-Finzi, 1969) or anthropogenic activities (Butzer, 1974).

4 Colluvial sections in subhumid landscapes (including southern Africa) show complex consequences of deposition, stability and incision (Watson et al., 1984). Ongoing luminescence dating is beginning to give an indication of the complexity of chronology.

5 Terminal lake basins (pans, playas, etc.) can respond dramatically, in terms of both their extent and the rapidity of change that can occur, to episodic rainfall events in their catchments. The history of Lake Eyre in Australia in the twentieth century bears witness to this fact (Figure 12.1).

6 West African dust-storm activity has shown very marked shifts in the past few decades in response to runs of dry years and increasing land-use pressures (Goudie and Middleton, 1992). The data for Nouakchott (Mauritania) are especially instructive, revealing a sudden acceleration in dust-storm events since the 1960s. A broadly similar picture could be presented for the High Plains of the USA during the 'Dust Bowl years' of the 1930s.

7 Some dune fields have also proved to be prone to repeated fluctuations in deposition and stabilization in the Holocene, and as more ^{14}C and luminescence dates become available the situation is likely to prove to be even more complex than hitherto believed. A

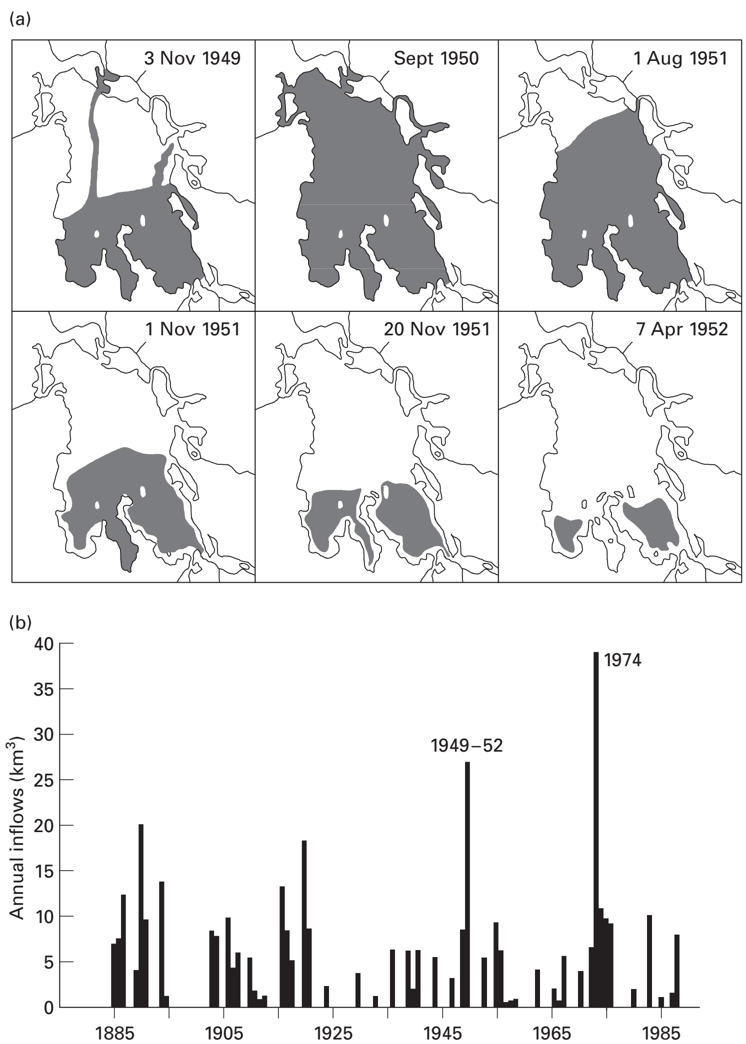

Figure 12.1 The flooding of Lake Eyre. (a) Extent of flooding in 1949–1952 (after Bonython and Mason in Mabbutt, 1977, figure 54). (b) Estimated annual inflows to Lake Eyre North for the period 1885–1989 (based on Kotwicki and Isdale, 1991, figure 2). Both 1949–1952 and 1974 were strong La Niña events. (Source: Viles and Goudie, *Earth-science Reviews*, 2003, volume 61: 105–131.)

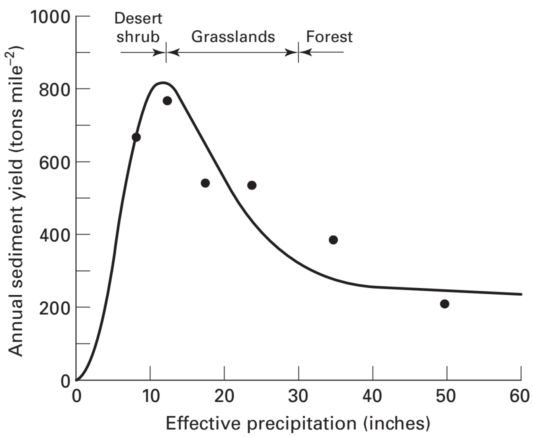

Figure 12.2 Variation of sediment yield with climate as based on data from small watersheds in the USA (after Langbein and Schumm, 1958).

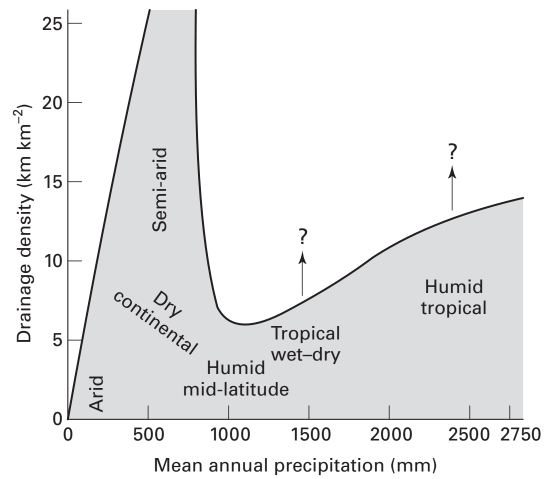

Figure 12.3 Relation between drainage density and mean annual precipitation. (After Gregory, 1976. Copyright © 1976. Reprinted by permission of John Wiley and Sons, Ltd.)

good illustration of this is provided by Gaylord's (1990) work in the Clear Creek area of south-central Wyoming where over the past 7500 years at least four episodes of enhanced eolian activity and aridity are recorded.

8 Various studies have shown how rates of denudation (e.g., Langbein and Schumm, 1958) and drainage density can change very rapidly in semi-arid areas either side of a critical rainfall or precipitation/evapotranspiration (P/E) value related to vegetation cover that constitutes a particular threshold between equilibrium states (Figures 12.2 and 12.3).

This apparent instability and threshold dependence of a range of arid zone landforms, rates, and processes leads one to believe that such areas may be especially susceptible to the effects of global warming, should this occur in the coming decades (Figure 12.4).

Climate changes in the past

Past climatic changes have had dramatic effects on deserts (Goudie, 2002) and this makes it likely that future climatic changes will also have dramatic effects. For example, climatic changes in the Holocene have led to the expansion and contraction of lakes.

In the Sahara there are huge numbers of Holocene pluvial lakes both in the Chotts of Tunisia and Algeria, in Mali (Petit-Marie et al., 1999) and in the south (e.g., Mega-Chad). In the Western Desert of Egypt and the Sudan, there are many closed depression or playas, relict river systems, and abundant evidence of prehistoric human activity (Hoelzmann et al., 2001). Playa sediments indicate that they once contained substantial bodies of water, which attracted Neolithic settlers. Many of these sediments have now been radiocarbon dated and indicate the ubiquity of an early to mid-Holocene wet phase, which has often been termed the 'Neolithic Pluvial'. A large lake, 'The West Nubian Palaeolake', formed in the far northwest of Sudan, (Hoelzmann et al., 2001). It was especially extensive between 9500 and 4000 years BP, and may have covered as much as 7000 km^2. If it was indeed that big, then a large amount of precipitation would have been needed to maintain it – possibly as much as 900 mm compared with the less than 15 mm it receives today. The Sahara may have largely disappeared during what has been called 'The Greening of Africa'.

Conversely, some deserts experienced abrupt and relatively brief drought episodes which caused extensive dune mobilization in areas such as the American High Plains (Arbogast, 1996).

At the present day, deserts continue to change, sometimes as a result of climatic fluctuations such as the run of droughts that has afflicted the West African Sahel over recent decades. The changing vegetation conditions in the Sahel, which have shown marked

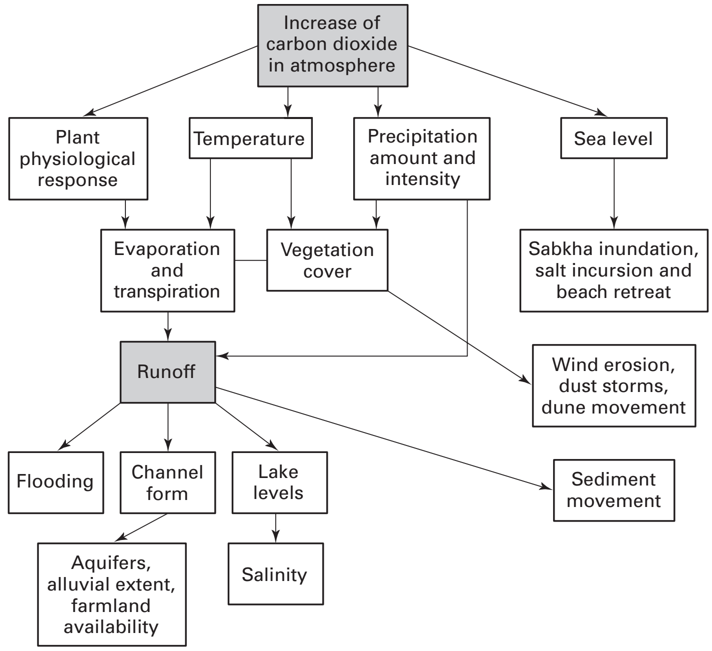

Figure 12.4 The potential responses of desert environments to increases of carbon dioxide in the atmosphere.

fluctuations since the mid-1960s, have been mapped by Tucker et al. (1991). Interannual variability in the position of the southern boundary of the Sahara, as represented by the 200 mm isohyet, can be explained in large measure by changes in the North Atlantic Oscillation and the El Niño–Southern Oscillation (ENSO) (Oba et al., 2001). The drought also led to reductions in the flow of rivers such as the Senegal and Niger, and to an increase in dust-storm activity (Middleton, 1985). The extent of Lake Chad's water surface has also fluctuated dramatically (Nicholson, 1996). It was particularly badly affected by the drought, which caused the lake surface area to fall from *c.* 25,000 km^2 in 1960 to just a tenth of that figure in the mid-1980s (Birkett, 2000).

We also know from recent centuries that ENSO fluctuations have had important consequences for such phenomena as droughts and dune reactivation (Forman et al., 2001), and through their effect on vegetation cover have had major impacts on slope stability and soil erosion (Viles and Goudie, 2003). Lake levels have responded to El Niño influences (Arpe et al., 2000), large floods have occurred, valleys have been trenched (Bull, 1997), erosivity patterns have altered (D'Odorico et al., 2001), and landslides and debris flows have been generated (Grosjean et al., 1997).

Wind erosion of soils

Changes in climate could affect wind erosion either through their impact on erosivity or through their effect on erodibility.

Erosivity is controlled by a range of wind variables including velocity, frequency, duration, magnitude, shear, and turbulence. Such wind characteristics vary over a whole range of timescales from seconds to millennia. For example, Bullard et al. (1996) have shown how dune activity varies in the South West Kalahari in response to decadal-scale variability in wind velocity, while over a longer timescale there is evidence that trade-wind velocities may have been elevated during the Pleistocene glacials. Unfortunately, general circulation models as yet give little indication of how wind characteristics might be modified in a warmer world, so that prediction of future changes in wind erosivity is problematic.

Erodibility is largely controlled by vegetation cover and surface type, and both of these can be influenced markedly by climatic conditions. In general, vegetation cover, which serves to protect the ground surface and to modify the wind regime, decreases as conditions become more arid. Likewise climate affects the nature of surface materials by controlling their moisture content, the nature and amount of clay mineral content (cohesiveness), and organic levels. Soils that are dry have a low clay content and little binding humus and are highly susceptible to wind erosion.

However, modeling the response of wind erosion to climatic variables on agricultural land is vastly complex, not least because of the variability of soil characteristics, topographic variation, the state of plant growth and residue decomposition, and the existence of windbreaks. To this needs to be added the temporal variability of eolian processes and moisture condition and the effects of different land management practices (Leys, 1999), which may themselves change with climate change.

Dust-storm activity

The changes in temperature and precipitation conditions that occurred in the twentieth century (combined with land-cover changes) had an influence on the development of dust storms (Goudie and Middleton, 1992). These are events in which visibility is reduced to less than 1 km as a result of particulate matter, such as valuable topsoil, being entrained by wind. This is a process that is most likely to happen when there are high winds and large soil-moisture deficits. Probably the greatest incidence of dust storms occurs when climatic conditions and human pressures combine to make surfaces susceptible to wind attack.

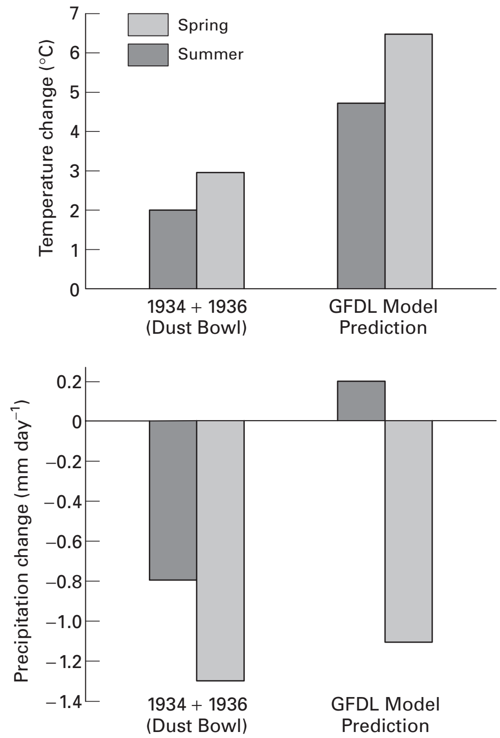

Figure 12.5 Comparison between two 'Dust Bowl years' (1934 and 1936) and the Geophysical Fluid Dynamic Laboratory (GFDL) model prediction for a $\times 2$ CO_2 situation for the Great Plains (Kansas and Nebraska) (modified from Smith and Tirpak, 1990, figure 7.3).

Possibly the most famous case of soil erosion by deflation was the Dust Bowl of the 1930s in the USA (see Chapter 4). In part this was caused by a series of hot, dry years which depleted the vegetation cover and made the soils dry enough to be susceptible to wind erosion, but the effects of this drought were gravely exacerbated by years of overgrazing and unsatisfactory farming techniques.

Attempts to relate past dust-storm frequencies to simple climatic parameters or antecedent moisture conditions have frequently demonstrated rather weak relationships (Bach et al., 1996), confirming the view that complex combinations of processes control dust emissions. Nevertheless, evidence is now emerging that relates dust emissions from Africa to changes in the North Atlantic Oscillation (Moulin et al., 1997).

If, however, soil moisture levels decline as a result of changes in precipitation and/or temperature, there is the possibility that dust-storm activity could increase in a warmer world (Wheaton, 1990). A comparison between the Dust Bowl years of the 1930s and model predictions of precipitation and temperature for the Great Plains of Kansas and Nebraska (Figure 12.5) indicates that mean conditions could be similar to those of the 1930s under enhanced greenhouse conditions (Smith and Tirpak, 1990), or even worse (Rosenzweig and Hillel, 1993).

If dust-storm activity were to increase as a response to global warming it is possible that this could have a feedback effect on precipitation that would lead to further decreases in soil moisture (Tegen et al., 1996; Miller and Tegen, 1998). It is also possible that increased dust delivery to the oceans could affect biogeochemical cycling, by providing nutrients to plankton, which in turn could draw down carbon dioxide from the atmosphere (Ridgwell, 2002). However, the impact and occurrence of dust storms will depend a great deal on land management practices, and recent decreases in dust-storm activity in North Dakota have resulted from conservation measures (Todhunter and Chihacek, 1999).

Sand dunes

Sand dunes, because of the crucial relationships between vegetation cover and sand movement, are highly susceptible to the effects of changes of climate (Figure 12.6). Some areas, such as the South West Kalahari (Stokes et al., 1997) or portions of the High Plains of the USA (Gaylord, 1990), may have been especially prone to the effects of changes in precipitation and/or wind velocity because of their location in climatic zones that are close to a climatic threshold between dune stability and activity.

One of the more remarkable discoveries of recent years, brought about by the explosive development in the use of thermoluminescent and optical dating of sand grains and studies of explorers' accounts (Muhs and Holliday, 1996), is the realization that such marginal dune fields have undergone repeated phases of change at decadal and century timescales in response to extended drought events during the course of the Holocene. Dates for reaction phases are given for the Nebraskan Sandhills by Stokes and Swinehart (1997) and Muhs et al. (1997), for Kansas by Arbogast (1996), and for the South West Kalahari by Thomas et al. (1997) (Figure 12.7).

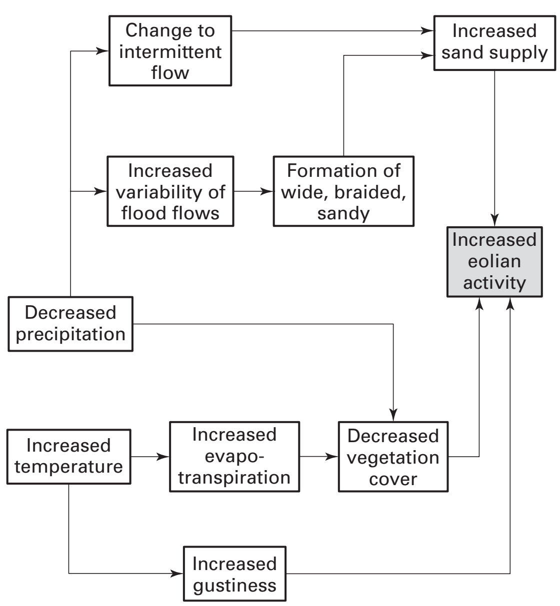

Figure 12.6 The influence of decreased precipitation and increased temperatures on eolian activity.

The mobility of desert dunes (M) is directly proportional to the sand-moving power of the wind, but indirectly proportional to their vegetation cover (Lancaster, 1995: 238). An index of the wind's sand-moving power is given by the percentage of the time (W) the wind blows above the threshold velocity (4.5 m per second) for sand transport. Vegetation cover is a function of the ratio between annual rainfall (P) and potential evapotranspiration (PE). Thus, $M = W/(P/PE)$. Empirical observations in the USA and southern Africa indicate that dunes are completely stabilized by vegetation when M is < 50, and are fully active when M is > 200.

Muhs and Maat (1993) have used the output from general circulation models (GCMs) combined with this dune mobility index to show that sand dunes and sand sheets on the Great Plains are likely to become reactivated over a significant part of the region, particularly if the frequencies of wind speeds above the threshold velocity were to increase by even a moderate amount. However, the methods used to estimate future dune field mobility are still full of problems and much more research is needed before we can have confidence in them (Knight et al., 2004).

For another part of the USA, Washington State, Stetler and Gaylord (1996) have suggested that with a 4°C warming vegetation would be greatly reduced and that as a consequence sand dune mobility would increase by over 400%.

For the Canadian Prairies, Wolfe (1997) found that while most dunes were currently inactive or only had active crests, under conditions of increased drought (as occurred in 1988 and are likely to occur in the future) most dunes would become more active (Figure 12.8).

The consequences of dune encroachment and reactivation could be serious and might lead to a loss of agricultural land, the overwhelming of buildings, roads, canals, runways, and the like, abrasion of structures and equipment, abrasion of crops, and the impoverishment of soil structure.

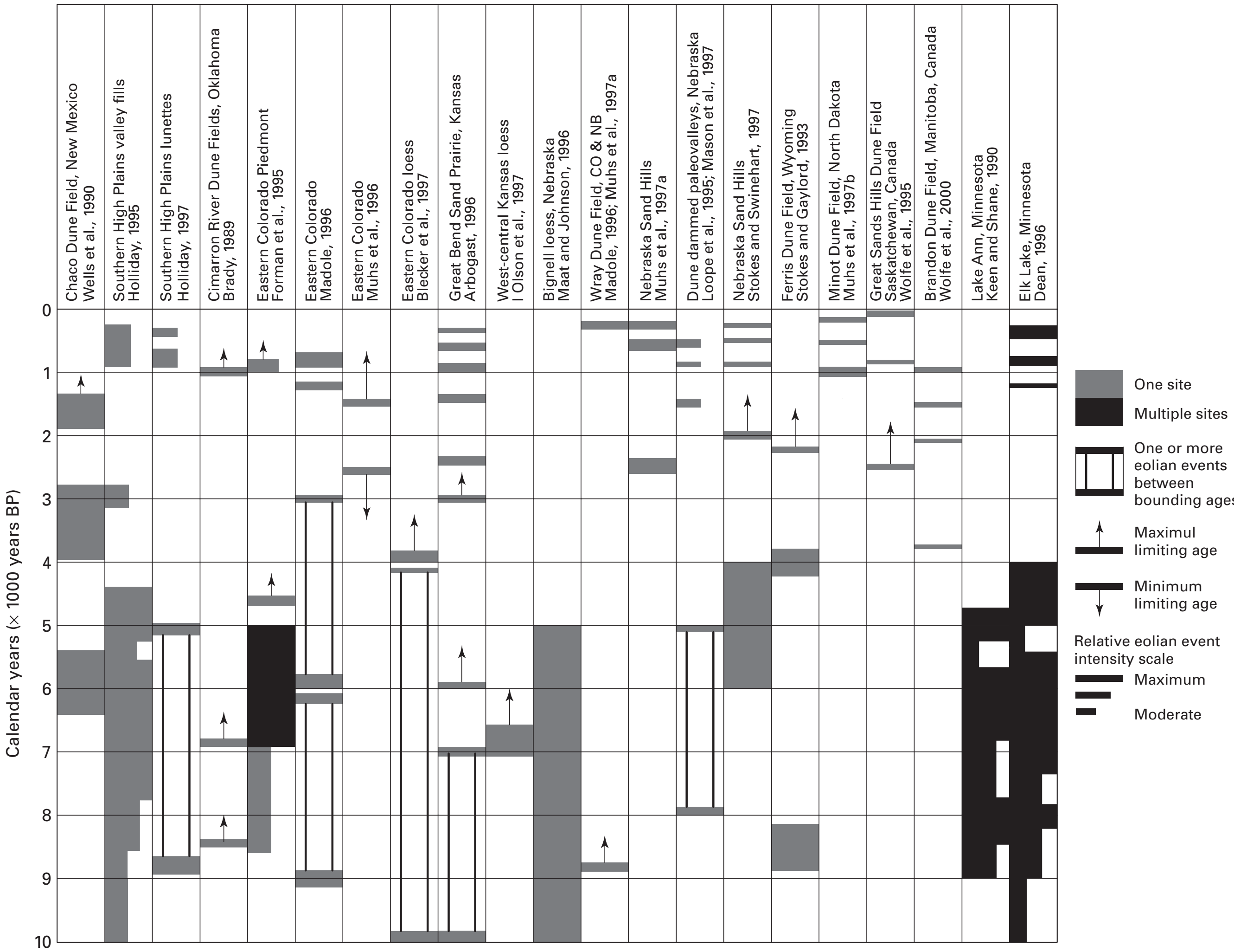

Figure 12.7 Summary of dune field activity and associated loess deposition for the Great Plains, USA. The terrestrial eolian record is also compared with the eolian input record from Lake Ann and Elk Lake, Minnesota. Note that there is evidence for sustained aridity between 10,000 and 5000 years ago and numerous discrete events in the past 2000 years. Length of solid bar reflects duration of eolian events and inferred dating errors. (Source: S. L. Forman et al., 2001, figure 13.)

Rainfall and runoff

A recent attempt to estimate the effects of global warming on runoff is provided by the UK Meteorological Office (Arnell, 1999a). What is clear from this work is that there will be clear differences at a global scale, with some areas generating more runoff and some generating less. However, with respect to dry regions, some of these will suffer particularly large diminutions in annual runoff (sometimes 60% or more). They appear to be more vulnerable than humid regions (Guo et al., 2002).

The sensitivity of runoff to changes in precipitation is complex, but in some environments quite small changes in rainfall can cause proportionally larger changes in runoff. This has been indicated for two areas of southern Australia by Chiew et al. (1996: 341):

In the south-west coast and the South Australian Gulf about 70% of the annual rainfall of 500 to 1000 mm occurs in the winter-half of the year. The streams in these regions generally flow for only 50% of the time, and on average less than 10% of the annual rainfall becomes runoff. It is also common for the total annual runoff to come from only one or

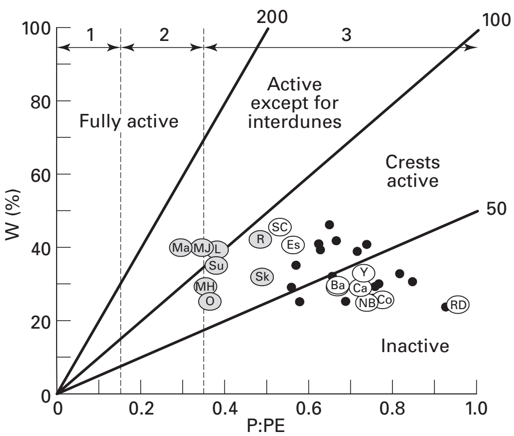

Figure 12.8 Dune mobility for stations across the Canadian Prairies for the 1988 drought year. Black dots represent values for 1961–1990 normals. Dune activity classes of Muhs and Holliday (1995) are also shown: (1) fully active; (2) largely active; (3) largely inactive. The names of the localities are Brandon (Ba), Broadview (Bv), Calgary (Ca), Coronation (Co), Estevan (Es), Lethbridge (L), Mayberries (Ma), Medicine Hat (MH), Moose Jaw (MJ), North Battleford (NB), Outlook (O), Red Deer (RD), Regina (R), Saskatoon (SK), Suffield (SU), Swift Current (SC), Yorkton (Y). Shaded stations are from drier, subhumid locations. (Source: Wolfe, 1997, figure 8.)

two significant flow events during winter. The simulations indicate that the average annual runoff increases at a much faster rate than the corresponding increase in rainfall. A rainfall increase of 10% enhances runoff by 50%, an increase of 20% more than doubles runoff, and an increase of 40% results in runoff being almost four times greater. A decrease in rainfall has a potentially more serious consequence as the amount of streamflow drops very quickly. A decrease in rainfall by 20% reduces runoff by one third while a decrease of 40% reduces runoff by 90%. The soil wetness in the winter-half changes at almost twice the rate of the change in rainfall for changes of rainfall of up to 20%. The changes, however, approach lower and upper limits with the simulations indicating that it is unlikely for the soil wetness to drop below 10% or increase above 70% even for rainfall changes of more than 50%.

Highly significant runoff changes may also be anticipated for semi-arid environments, such as the southwest USA. The models of Revelle and Waggoner (1983) suggest that the effects of increased evapotranspiration losses as a result of a 2°C rise in temperature would be particularly serious in those regions where the mean annual precipitation is less than about 400 mm (Table 12.1). Projected summer dryness in

Table 12.1 Approximate percentage decrease in runoff for a 2°C increase in temperatures. Source: from data in Revelle and Waggoner (1983)

Initial mean temperature (°C)	*Precipitation (mm year^{-1})*					
	200	*300*	*400*	*500*	*600*	*700*
−2	26	20	19	17	17	14
0	30	23	23	19	17	16
2	39	30	24	19	17	16
4	47	35	25	20	17	16
6	100	35	30	21	17	16
8		53	31	22	20	16
10		100	34	22	22	16
12			47	32	22	19
14			100	38	23	19

such areas may be accentuated by a positive feedback process involving decreases in cloud cover and associated increases in radiation absorption on the ground consequent upon a reduction in soil moisture levels (Manabe and Wetherald, 1986). Our modeling capability in this area is still imperfect and different types of model indicate differing degrees of sensitivity to climatic change (Nash and Gleick, 1991).

Shiklomanov (1999) has suggested that in arid and semi-arid areas an increase in mean annual temperature by 1° to 2°C and a 10% decrease in precipitation could reduce annual river runoff by up to 40–70% (Table 12.2).

One factor that makes estimates of rainfall–runoff relationships complicated is the possible effect of higher CO_2 concentrations on plant physiology and transpiration capacity. At higher CO_2 concentrations transpiration rates are less and this could lead to an increase in runoff or at least limit its reduction (e.g., Idso and Brazel, 1984). It is also important to remember that future runoff will be conditioned by nonclimatic factors, such as land-use and land-cover change, the construction of reservoirs, groundwater storage, and water demand (Conway et al., 1996).

River channels

Channels in arid regions are particularly sensitive to changes in precipitation characteristics and runoff (Nanson and Tooth, 1999). They can display rapid changes between incision and aggradation over short

Table 12.2 Results of assessments of impacts of climate change on annual river runoff (basins and areas of water deficit). Source: after various sources in Shiklomanov (1999, table 2.1)

Region and basin	*Temperature*	*Precipitation*	*Change in annual runoff (%)*
Mean for seven large basins in the western USA	+2	–10	–40 to –76
Colorado River, USA	+1	–10	–50
Peace River, USA	+1	–10	–50
River basins in Utah and Nevada, USA	+2	–10	–60
River basins in the steppe zones of European Russia	+1	–10	–60

time periods in response to quite modest changes in climate. This is particularly true in the case of the arroyos of the American southwest (Cooke and Reeves, 1976; Balling and Wells, 1990), which have displayed major changes in form since the 1880s (see Chapter 6). There has been considerable debate as to the causes of phases of trenching, and it is far from easy to disentangle anthropogenic from climatic causes, but in many cases it is fluctuations in either rainfall amount or intensity that have been the controlling factor (Hereford, 1984; Graf et al., 1991). Thus, the sorts of changes in runoff discussed in the last section could have a profound influence on channel characteristics. These in turn have an impact on humans because they can lead to changes in the agricultural suitability of bottomlands, modify local aquifers, modify sediment inputs into reservoirs, and cut into engineering structures. Arroyo incision can also lead to the draining of riverbed marshes (cienegas). It may even have produced settlement abandonment (Hereford et al., 1995).

Lake levels

Closed depressions are a widespread phenomenon in arid lands and their water levels and salinity characteristics respond rapidly and profoundly to climatic changes (Grimm et al., 1997). This generalization applies both to large and small lakes. In the twentieth century, for example, some of the largest arid zone lakes (e.g., Chad, the Aral Sea, the Caspian, and the Great Salt Lake of Utah) have shown large variations in their extents, partly in some cases because of human activities, but also because of climatic fluctuations within their catchments. They are sensitive to climate variability that would be of minor influence in systems with outflows.

For example, from the early 1950s to the middle 1980s the total area of lakes in China with an individual area of over 1 km^2 declined from 2800 to 2300 km^2 and the whole area of China's lakes has been reduced from 80,600 km^2 to 70,988 km^2 (Liu and Fu, 1996). An increasingly warm and dry climate was the principal cause of the reduced lake area on the Qingzang Plateau, Northwest China, the Inner Mongolian Plateau, and the North China Plain.

Likewise, in the early 1960s, prior to the development of the Sahel Drought, Lake Chad had an area of 23,500 km^2, but by the 1980s had split into two separate basins and had an area of only 1500 km^2. The Caspian Sea was –29.10 m in 1977, but in 1995 had risen to –26.65 m, an increase of 2.45 m in just 17 years. Similarly impressive changes have occurred in recent decades in the level of the Great Salt Lake in Utah, with a particularly rapid rise taking place between 1964 and 1985 of nearly 20 feet (6 m).

Changes in lake level of these sorts of magnitude have an impact on a diverse range of human activities, ranging from fisheries and irrigation to recreation. Moreover, the drying-up of lake beds can have adverse effects on air quality and human health through the liberation of dust, as has been found as a result of the humanly induced desiccation of the Aral Sea and Owens Lake (Reheis, 1997).

Sea-level rise and arid-zone coastlines

One consequence of global warming will be a rise in sea level at around 5 mm per year (50 cm per century). Arid coastlines could be affected by this process. This could potentially be especially serious for low-lying coastlines such as the Sabkhas of the United Arab

Emirates and elsewhere in the Middle East. Sabkhas result from the interaction of various depositional and erosional processes that create a low-angle surface in the zone of tidal influence. This means that they are subject to periodic inundation and might be vulnerable to modest sea-level rise and to any increase in storm-surge events. Given the degree of infrastructural development that has taken place in their proximity, this is a serious issue for cities such as Abu Dhabi. However, it is likely that many sabkhas will be able to cope with modestly rising sea levels, for a range of processes contribute to their accretion. These include algal stromatolite growth, fecal pellet deposition, eolian inputs, and evaporite precipitation. Some of these can cause markedly rapid accretion (Schrieber, 1986), even in the absence of a very well developed plant cover. Moreover, as sea level (and groundwater) rises, surface lowering by deflational processes will be reduced. An example of a sabkha that may well maintain itself, or even continue to aggrade in spite of sea-level rise, is provided by the Umm Said Sabkha of Qatar, where eolian dune inputs from inland cause the sabkha to build out into the Arabian Gulf.

Among arid-zone coastal environments that may be particularly susceptible to sea-level rise are deltaic areas subject to subsidence and sediment starvation (e.g., the Nile) and areas where ground subsidence is occurring as a result of fluid abstraction (e.g., California). Whereas the IPCC prediction of sea-level rise is 30–100 cm per 100 years, rates of deltaic subsidence in the Nile Valley are 35–50 cm per 100 years, and in other parts of the world rates of land subsidence produced by oil, gas, or groundwater abstraction can be up to 500 cm per 100 years.

Rising sea levels can be expected to cause increased flooding, accelerated erosion, and accelerated incursion of saline water up estuaries and into aquifers. Coastal lagoon, spit, and barrier systems (such as those of Ras Al Khaimah) may be especially sensitive (Goudie et al., 2000), as will coastlines that have been deprived of sediment nourishment by dam construction across rivers.

Salt weathering

Salt weathering is a major geomorphologic process in many drylands and it poses a threat to many man-made structures, both ancient and modern (Goudie and Viles, 1997).

The most cited cause of salt weathering is generally the process of salt crystal growth from solutions in rock pores and cracks (Evans, 1970). Another type of salt weathering process is salt hydration. Certain common salts hydrate and dehydrate relatively easily in response to changes in temperature and humidity. As a change of phase takes place to the hydrated form, water is absorbed. This increases the volume of the salt and thus develops pressure against pore walls. Sodium carbonate and sodium sulfate both undergo a volume change in excess of 300% as they hydrate.

A third possible mechanism of rock disruption through salt action has been proposed by Cooke and Smalley (1968), who argue that disruption of rock may occur because certain salts have higher coefficients of expansion than do the minerals of the rocks in whose pores they occur.

In addition to these three main categories of mechanical effects, salt can cause chemical weathering. Some saline solutions can have elevated pH levels. Why this is significant is that silica mobility tends to be greatly increased at pH values greater than 9. Indeed, according to various studies, silica solubility increases exponentially above pH 9. The presence of sodium chloride may also affect the degree and velocity of quartz solution. At higher NaCl concentrations quartz solubility and the reaction velocity both increase. The growth of salt crystals may be able to cause pressure solution of silicate grains in rocks, for silica solubility increases as pressure is applied to silicate grains. This is a mechanism that has been identified as important in areas where calcite crystals grow, as for example in areas of calcrete formation.

The attraction of moisture into the pores of rocks or concrete by hygroscopic salts (e.g., sodium chloride) can accelerate the operation of chemical weathering processes and of frost action (MacInnis and Whiting, 1979) and the disruptive action of moisture trapped in rock capillaries is well known.

Salt can have a deleterious impact on iron and concrete. Many engineering structures are made of concrete containing iron reinforcements. The formation of the corrosion products of iron (i.e., rust) causes a volume expansion to occur. If one assumes that the prime composition of such corrosion products is $Fe(OH)_3$, then the volume increase over the uncorroded iron

Table 12.3 Global warming and salinization

Negative effects	*Positive effect*
Increased moisture stress will lead to a greater need for irrigation	Groundwater levels may fall
Higher evaporation losses will lead to increased salt concentrations	
Lesser freshwater inputs will reduce estuarine flushing and decrease replenishment in coastal aquifers	
Higher sea levels will cause change of salinity in estuaries and in coastal aquifers	

can be fourfold. Thus when rust is formed on the iron reinforcements, pressure is exerted on the surrounding concrete. This may cause the concrete cover over the reinforcement to crack, which in turn permits the ingress of oxygen and moisture, which then aggravates the corrosion process. In due course, spalling of concrete takes place, the reinforcements become progressively weaker, and the whole structure may suffer deterioration.

Sulfates can cause severe damage to, and even complete deterioration of, Portland cement concrete. Although there is controversy as to the exact mechanism of sulfate attack (Cabrera and Plowman, 1988), it is generally accepted that the sulfates react with the alumina-bearing phases of the hydrated cement to give a high sulfate form of calcium aluminate known as ettringite. The formation of ettringite involves an increase in the volume of the reacting solids, a pressure build up, expansion and, in the most severe cases, cracking and deterioration. The volume change on formation of ettringite is very large, and is even greater than that produced by the hydration of sodium sulfate.

Another mineral formed by sulfates coming into contact with cement is thaumasite. This causes both expansion and softening of cement and has been seen as a cause of disintegration of rendered brickwork and of concrete lining in tunnels.

Global warming could influence salt weathering in a variety of ways (Table 12.3). In low-lying coastal areas sea-level rise will have an impact upon groundwater conditions. The balance between freshwater aquifers and seawater incursion is a very delicate one. Recent modeling, using the Ghyben–Herzberg relationship, suggests that in the Nile Delta, where subsidence is occurring at a rate of 4.7 mm per year, overpumping is taking place, and less freshwater flushing is being achieved by the dammed Nile, a 50 cm rise in Mediterranean level will cause an additional intrusion of saltwater by 9.0 km into the Nile Delta aquifer (Sherif and Singh, 1999).

In addition, many arid areas may become still more arid. Moisture deficits will increase and streamflows will decrease, partly because of increased evaporative loss promoted by higher temperature, and partly because of decreased precipitation inputs. Lakes and reservoirs will suffer more evaporative losses and so could become more saline. In many areas streamflow will decline by 60–70%. Greater moisture stress may lead to an expansion in the quest for irrigation water, there may be less freshwater inputs to flush out estuarine zones, and rates of evaporative concentration of salt may increase (Utset and Borroto, 2001). Such tendencies could lead to an increase in salinization (Imeson and Emmer, 1992; Szabolcs, 1994). Conversely, in some areas falling groundwater levels may follow on from decreased recharge, and this could lead to improvements in the incidence of groundwater-induced salinization.

Points for review

Are desert areas prone to large and rapid geomorphologic and hydrologic changes in response to apparently modest climatic stimuli?

To what extent may wind erosion of soils, dust-storm activity and dune mobility change in a warmer world?

Guide to reading

Goudie, A. S., 2002, *Great warm deserts of the world: landscapes and evolution*. Oxford: Oxford University Press. A survey of the response of the world's deserts to past climatic changes.

Williams, M. A. J. and Balling, R. C. Jr., 1996, *Interactions of desertification and climate*. London: Arnold. A consideration of human impacts on desert environments, including a discussion of the effects of future climate changes.